RAINFALL IN MARITIME TROPICAL AIR OVER THE MIDWEST, JULY 16–18, 1953

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INTRODUCTION

The rainfall over the Midwest from July 16–18, 1953 aroused the curiosity of all interested in the relationship between weather and the weather map. The fronts that are often invoked to explain the weather were absent. When rain occurs without the aid of fronts, as in this case, the meteorologist must seek other explanations of the phenomenon. This article presents an investigation of the possible causes in this instance of such precipitation and an attempt to find a model suitable for portrayal on the weather map. It is concluded that a theoretical model proposed by Bjerknes and Holmboe seems suitable.

THE WEATHER AND THE ANALYSIS

The rain fell from a patch of thunderstorms that moved from Kansas into Ohio. Scattered thunderstorms over Kansas on July 15 had moved and formed into a general rain area over Missouri by the 16th. The rainfall continued during the morning of the 17th when the patch was over Illinois and Indiana, and in the next 24 hours it moved over Michigan and Ohio (see figs. 1–4).

The rain patch was roughly 180 miles wide and 480 miles long along the north-northwest south-southeast axis (fig. 3) by July 17, and maintained approximately this size and shape for the next 24 hours while the axis moved with a uniform speed. The rain area took about 24 hours to move over a given station with a total precipitation averaging about ½ inch and a maximum total of 1¾ inches. The cloud bases in the rain area were from 8,000 to 12,000 feet, except that where the precipitation was heaviest, they lowered to about 800 feet.

Concurrent with the beginning of the rain on July 15 a warm front along the Gulf Coast was dissipating as shown by the frontal analysis of the weather map (fig. 2), and maritime tropical air had penetrated northward to the Canadian border. A moist tongue over Oklahoma and Kansas (fig. 1) at this time moved eastward with the rain (fig. 3). Aloft, an old cut-off cold vortex evolved into an open trough which moved eastward behind the rain patch (fig. 5). The trough aloft had extended downward to the surface by 0330 GMT July 18 (fig. 4) but earlier the surface analysis had had no feature associated with the rain.

The radiosonde observation for Green Bay, Wis., at 1500 GMT July 16 (fig. 6) is representative of the air mass except where precipitation was occurring. Note that the pseudo-wet-bulb curve decreases more rapidly in the vertical, between 910 and 430 mb., than the moist adiabatic curve, indicating convective instability. Saturation of a convectively unstable air mass will produce convective cells [1] which result in the formation of convective type clouds, assuming that sufficient condensation nuclei are present. The convective clouds in this air mass became thunderstorms when their tops reached above the freezing level which was about 2,000 feet above the base of the clouds.

Saturation of a convectively unstable air mass is most easily effected by adiabatic lifting. The lift may be classified as (1) orographic lifting, (2) frontal lifting, or (3) vertical motion resulting from convergence in the wind field; the latter process of course would operate also in the first two. The effect of each will be examined as applied to this rain. (Hereafter, horizontal divergence will be referred to simply as divergence.)

A lift of 2,000 feet would have been necessary for saturation of that layer of air having the highest moisture content, which was the layer nearest the surface. Considering the trajectory of the air mass from the Gulf Coast to the Great Lakes region, the maximum orographic lifting could have been only about 800 feet, which would not have been sufficient, alone, to fulfill the above condition.

Frontal lifting was considered as a possible factor in the explanation of precipitation because of the presence of a warm front along the Gulf Coast at the beginning of the period and the possibility that this front had not in reality dissipated or that a new front had formed in the same general area (fig. 2). A convenient method for determining the existence of fronts is the examination of thickness charts for temperature discontinuities. Twelve hours prior to the dissipation of the front on the 0330 GMT surface chart, the maximum temperature gradient expressed in terms of 1,000–850 mb. thickness (fig. 8) showed a gradient of 50 feet per 180 miles measured northward and normal to the surface position; however, a gradient of similar magnitude existed south of the front thus revealing no temperature discontinuity. The thickness chart for 1500 GMT, July 17 (fig. 9), 36 hours later,

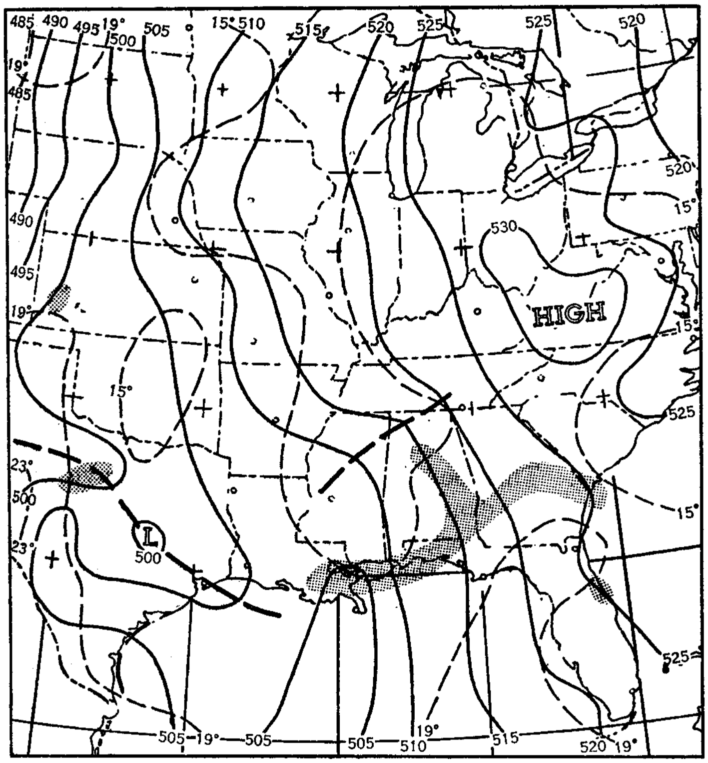

FIGURE 1.—850-mb. chart for 1500 GMT, July 15, 1953. Height contours (solid lines) are labeled in tens of feet and drawn for 50-foot intervals. Dew-point isopleths (light dashed lines) drawn for intervals of 4° C. Shaded areas indicate precipitation at the surface at 1530 GMT. Positions of predominant troughs indicated as heavy dashed lines.

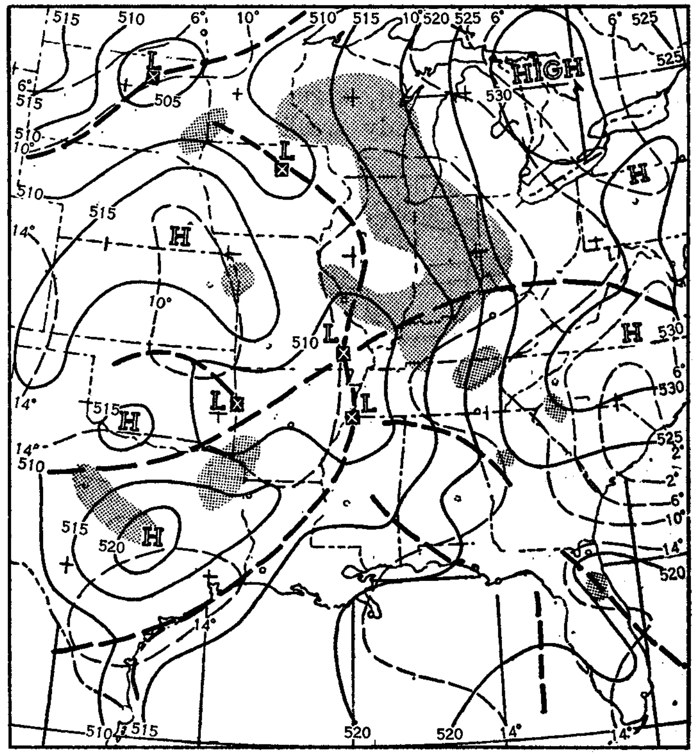

FIGURE 3.—850-mb. chart for 1500 GMT, July 17, 1953

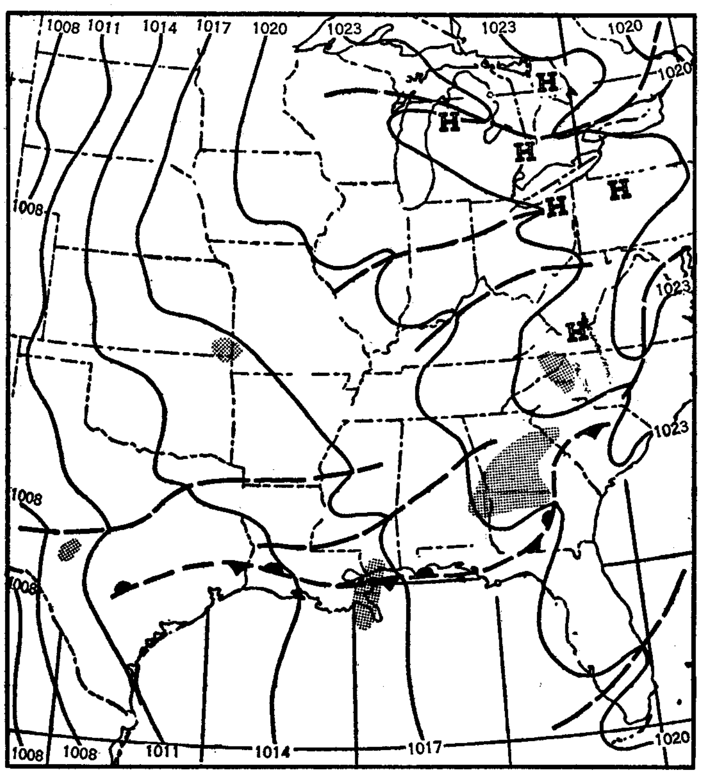

FIGURE 2.—Surface chart for 0330 GMT, July 16, 1953. Isobars (solid lines) drawn for 3-mb. intervals. Note the dissipating quasi-stationary front along the Gulf Coast area.

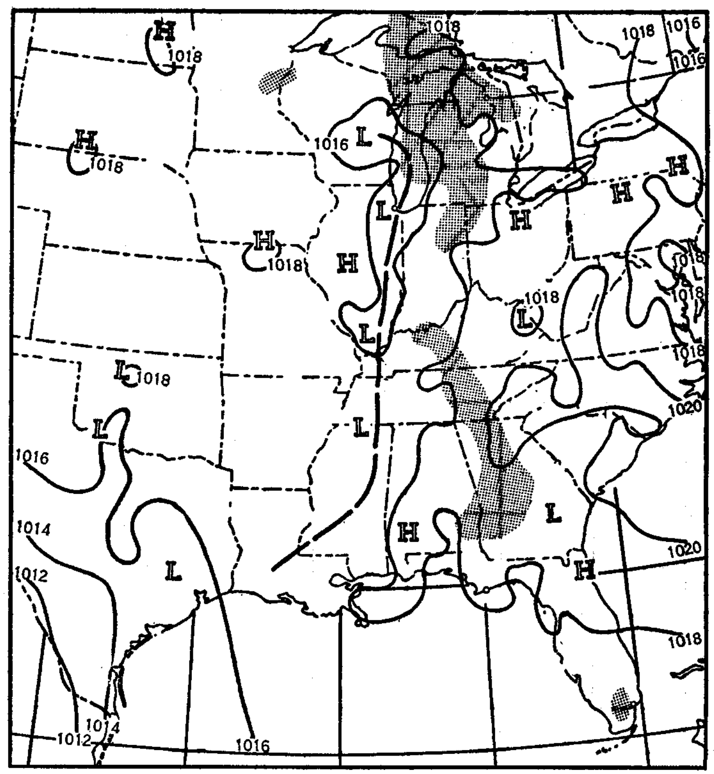

FIGURE 4.—Surface chart for 0330 GMT, July 18, 1953. Because of the extremely weak pressure gradient, isobars (here shown for 2-mb. intervals) were based on a detailed analysis using 1-mb. intervals with no attempt at smoothing.

shows a similarly weak and diffuse temperature field. Examination of thickness charts for higher layers for the same times (figs. 10 and 11) shows definite weakening of the temperature gradient, arguing for the progressive dissipation of the already weak front. Such a front would of necessity have been nearly vertical in slope. Any lifting over such a front would have produced precipitation only over Alabama and Mississippi since its lift would have been confined to the general area of the surface position. RAOBS from the Gulf Coast stations northward to the area of rain showed none of the characteristic changes of lapse rate normally associated with frontal surfaces.

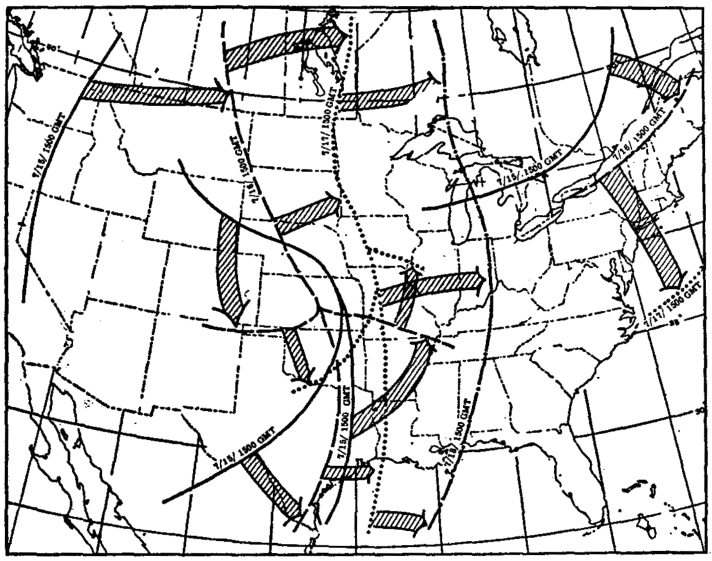

FIGURE 5.—24-hour 500-mb. trough positions from 1500 GMT, July 15, 1953, through 1500 GMT, July 18, 1953. The position at 1500 GMT, July 17, 1953 (the north-south dotted line) is that especially pertinent in its relation to the Bjerknes-Holmboe model referred to in the text.

In order to determine whether upward vertical velocities were present and of sufficient magnitude to effect the saturation of this convectively unstable air mass, a nomograph suggested by Bellamy [2] was used. Divergence for the standard constant pressure levels up to 300 mb. was calculated directly from the horizontal wind field for an area slightly larger than the rain patch at 1530 GMT July 17 (fig. 3). The good RAWIN coverage which gave winds in and above the clouds made these measurements possible. These values of divergence were then used to calculate the vertical velocities at the same constant pressure surfaces using the formulae and their graphical evaluation as developed by Kuhn [3].

Vertical velocity, w may be related to horizontal velocity divergence, div $\boldsymbol{V}$, by writing the equation of continuity in the form

$$\frac{\partial w}{\partial z}=-\frac{1}{\rho}\frac{d\rho}{dt}-\text{div}\,\boldsymbol{V}$$

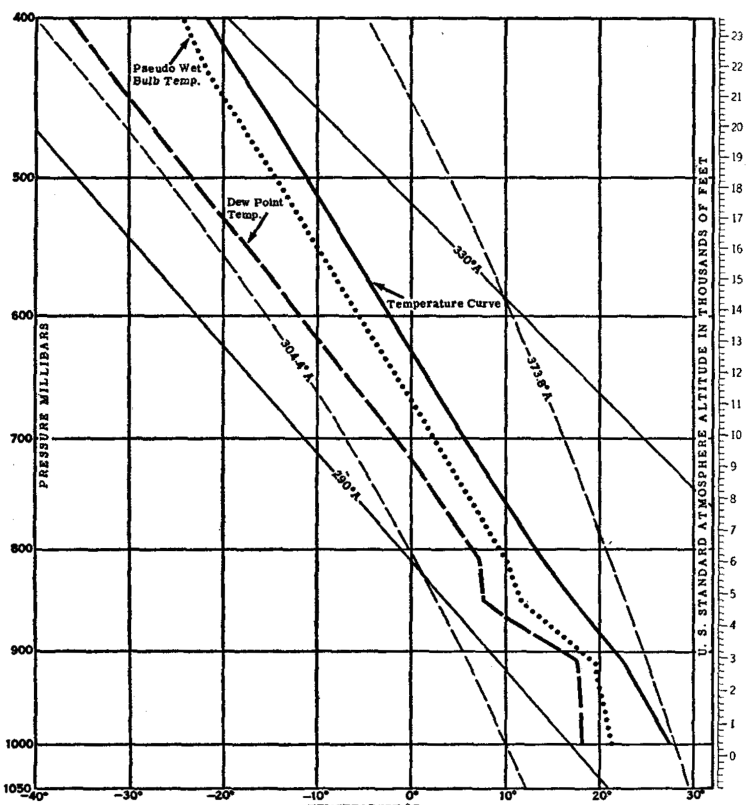

FIGURE 6.—Radiosonde observation from Green Bay, Wis. for 1500 GMT, July 16, 1953. Note that the lapse rate of the pseudo-wet bulb temperature curve is greater than or the same as the moist adiabatic lapse rate in the layer between 910 and 430 mb.

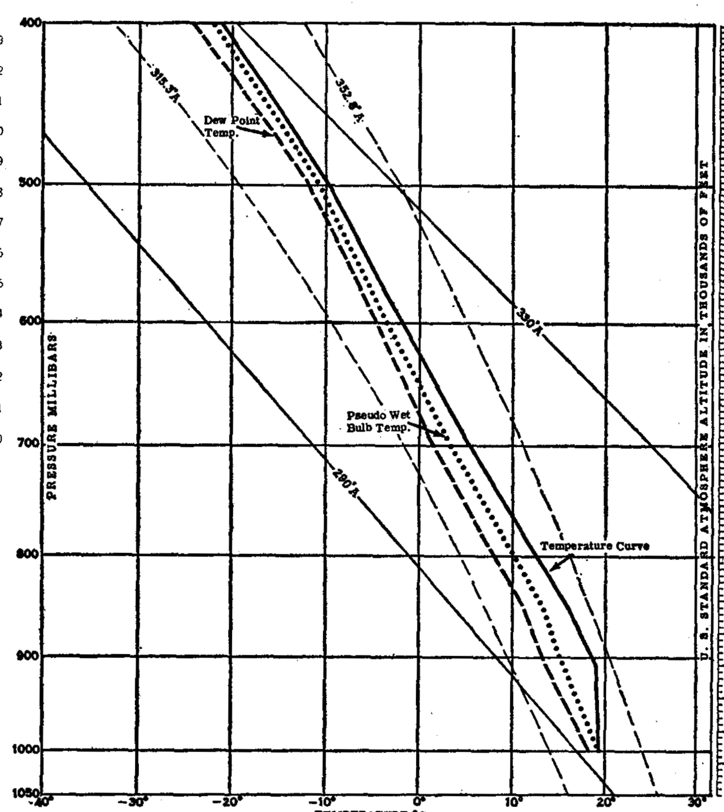

FIGURE 7.—Radiosonde observation from Green Bay, Wis. for 1500 GMT, July 17, 1953. Above 650-mb. the lapse rate of the temperature curve has decreased in 24 hours to slightly less than the moist adiabatic lapse rate.

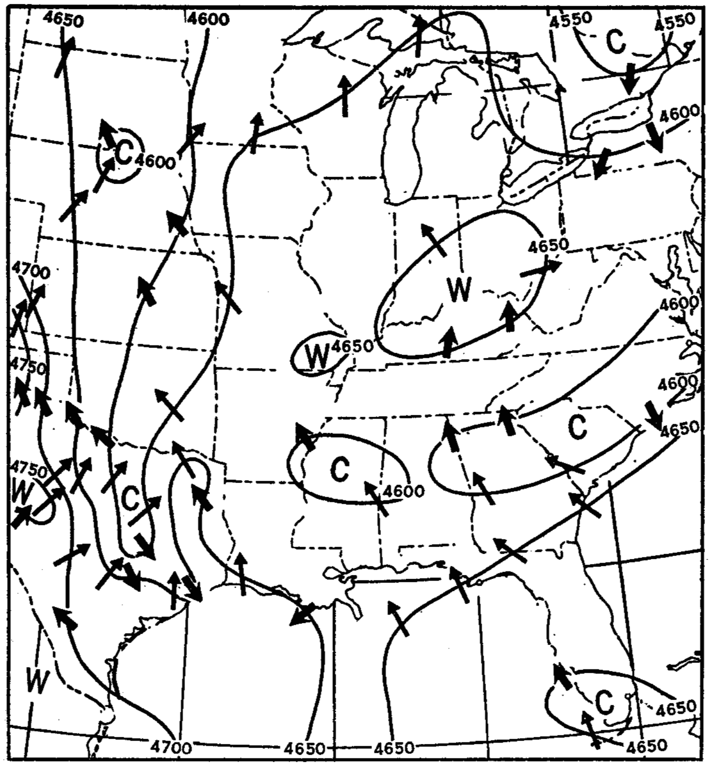

FIGURE 8.—1000-850-mb. layer thickness (or height difference) chart for 1500 GMT, July 15, 1953. Thickness isopleths are in feet for 50-foot intervals. Thin arrows indicate advection of warmer air and thick arrows advection of colder air. Note the generally weak gradients northward from the Gulf Coast.

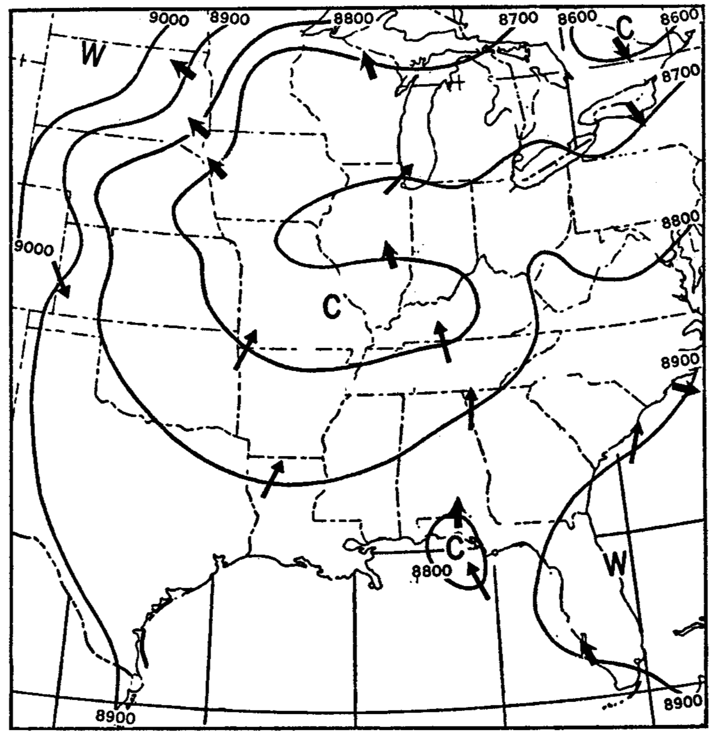

FIGURE 10.—700-500-mb. layer thickness chart for 1500 GMT, July 15, 1953. Thickness isopleths are in feet for 100-foot intervals.

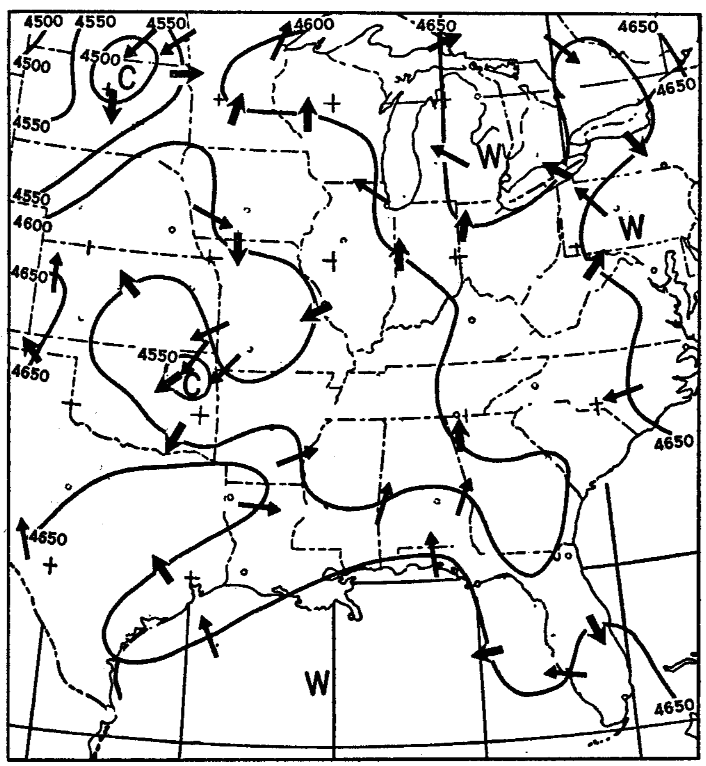

FIGURE 9.—1000-850-mb. layer thickness chart for 1500 GMT, July 17, 1953. Observe that over Illinois and Indiana, the general area of precipitation at this time (see fig. 3), the gradients are quite weak and some slight cold, rather than warm, advection is indicated.

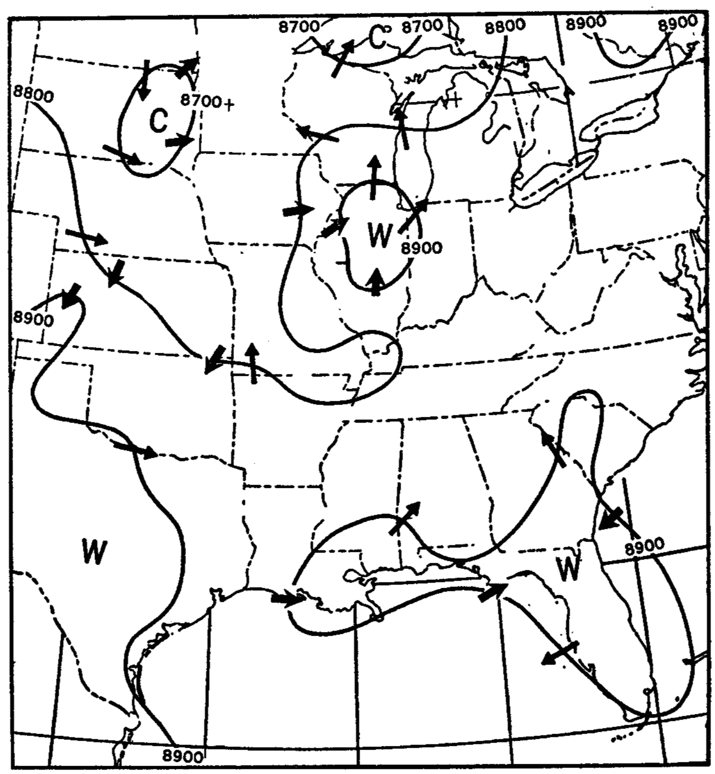

FIGURE 11.—700-500-mb. layer thickness chart for 1500 GMT, July 17, 1953.

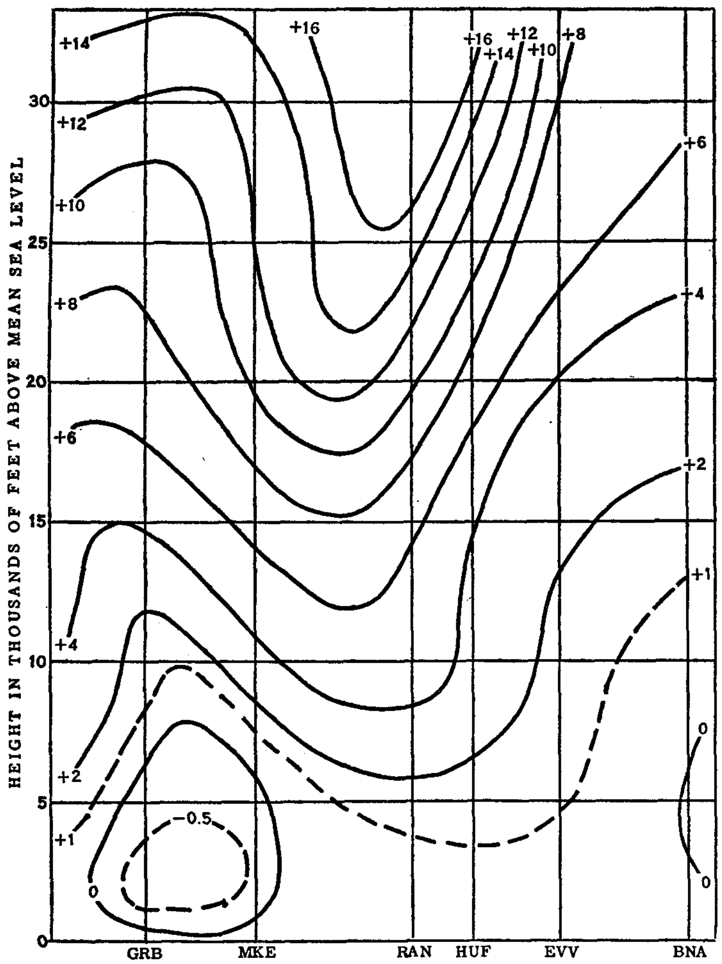

FIGURE 12.—Cross-section profile of vertical motion in the atmosphere from Green Bay, Wis., to Nashville, Tenn. (along an approximately straight line) for 1500 GMT, July 17, 1953. Isopleths are drawn for even intervals of every 2 cm. sec.$^{-1}$ (solid lines) and for odd intervals of +1 and −0.5 cm. sec.$^{-1}$. Plus values indicate upward velocities, zero indicates no vertical motion, and minus values represent downward motion. The data were calculated from divergence values at standard constant pressure surfaces after the method developed by Kuhn [3]. GRB=Green Bay, MKE=Milwaukee, Wis., RAN=Rantoul, Ill., EVV=Evansville, Ind., and BNA=Nashville.

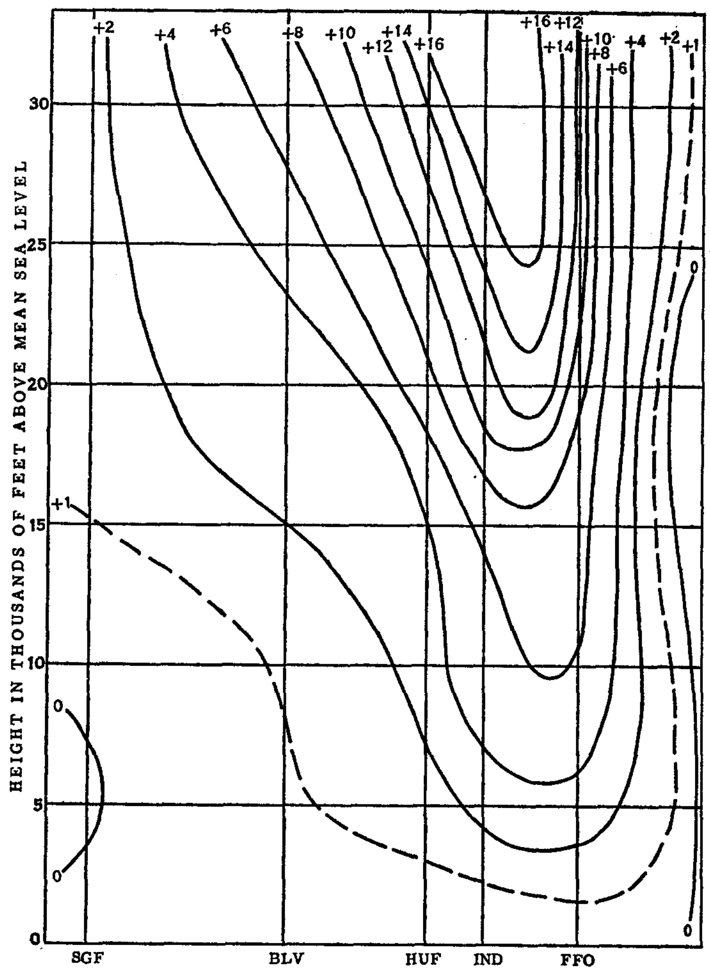

FIGURE 13.—Cross-section profile of vertical motion in the atmosphere from Springfield, Mo. to Dayton, Ohio for 1500 GMT, July 17, 1953. Note that this profile is not quite at right angles to the profile in figure 12; however, they intersect (at Terre Haute, Ind.) and picture two views of the same "cell" of maximum upward vertical velocity represented by the +16 isopleth. SGF=Springfield, BLV=Belleville, Ill., IND=Indianapolis, Ind., and FFO=Dayton.

where z is height, ρ is density, and t is time. The size of the term

$$\frac{1}{\rho}\frac{d\rho}{dt}$$

depends on the nonadiabatic effects and the adiabatic density change accompanying vertical motion. The nonadiabatic effects such as condensation, evaporation, and radiation from the cloud tops could not be estimated and were neglected. Austin [4] shows that in warm air masses moving northward, the nonadiabatic changes due to diurnal temperature variation and advective cooling may be ignored. Fleagle [5] found that for a magnitude of divergence of 10^{-5} sec^{-1} adiabatic density change is of a magnitude one order lower and may also be neglected. The neglect of these effects gives

$$\frac{\partial w}{\partial z} = -\text{div}\, \boldsymbol{V};$$

Kuhn's formulae for the constant pressure surfaces express this simple relationship. However, he approximated the density change by use of the Standard Atmosphere.

Isopleths of vertical velocity in cm sec^{-1} (upward velocity positive) were obtained by interpolation between the constant pressure surfaces. These are presented in two profiles (figs. 12 and 13), which are nearly perpendicular to each other and intersect approximately at the maximum value of upward vertical velocity. The continued increase of vertical velocity with height is due to two factors: continued convergence (−divergence) to above 500 mb. (see [6] for a similar vertical distribution of divergence), and the decrease of density with height [6]. The profiles show a "cell" of upward motion and suggest surrounding cells of downward motion. The north-south profile corresponds to the north-south axis of the instantaneous rain pattern for 1530 GMT July 17 (fig. 3). A comparison of the east-west profile with this rainfall pattern shows that the edges of the rain area coincide, roughly, with zero vertical velocities. Thus, upward vertical velocities coincided with precipitation. Such an apparent correlation has been found in many more cases by Panofsky [7]. He also shows that areas of few clouds and no precipitation correspond with areas of subsidence. Apparently, the lift shown in the profiles was preceded by downward vertical velocities.

According to Bjerknes' "slice" method for lifting a layer to saturation as presented by Petterssen [8], the greater the lapse rate, or the nearer the lapse rate is to the dry adiabatic slope, the greater the amount of convective cloudiness that may be expected. Subsidence prior to the time of precipitation and heat of condensation released above 10,000 feet are both processes which would decrease the lapse rate causing it to approach the moist adiabatic slope. Thus, by the "slice" method, the greater the effect of these two processes, the less the convective cloudiness that would be expected. Contrary to Petterssen's conclusion, it was observed in this case that the greatest convective cloudiness occurred when the lapse rate between 700 and 500 mb. approached the moist adiabatic slope (fig. 7). Empirical investigations by Austin [1] and others confirm this observation for many cases. Austin says, "For the development of the vertical accelerations to force the cloud upward it seems desirable that there be a steep lapse rate of temperature. However, . . . a condition which gives rise to a high liquid water content does not appear to favor the development of vertical accelerations. Therefore, it should not be expected that a steep lapse rate of temperature is necessarily the most favorable condition for cumulus growth."

Cressman [9] has treated the "slice" method in a manner similar to that of Bjerknes but replaces the assumption of no divergence with the assumption of divergence such that the inflow and outflow compensate each other. His analysis of this treatment shows that the effect of vertical velocities is greatest "when the actual lapse rate exceeds the moist adiabatic value only slightly." Cressman further states that in a region of upward motion the lapse rate reaches an equilibrium value, with a stabilizing tendency due to the condensation and precipitation being opposed by a destabilizing tendency due to the upward motion. Thus Cressman's work explains the lapse rates observed in the situation presented here.

The processes which stabilized the lapse rate in the rain area apparently counteracted the cold advection, shown on the 700–500 mb. thickness chart for 1500 GMT, July 15 (fig. 10), over southern Illinois that might have been expected to move northeastward with the movement of the cold trough. The chart 48 hours later (fig. 11) indicates an increase of 200 feet in thickness (or midtroposphere warming) over Illinois and Indiana at the time of the rain. Definitely the cold trough did not release the rain by midtroposphere cooling, but rather by the dynamic effect of divergence.

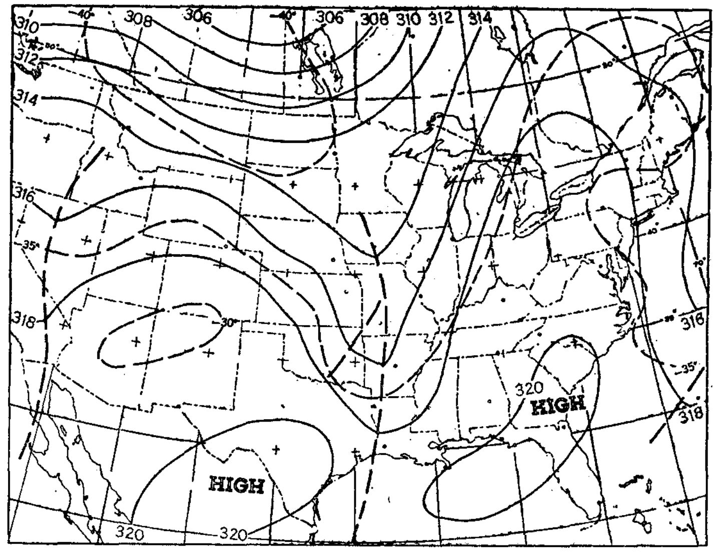

FIGURE 14.—300-mb. chart for 1500 GMT, July 17, 1953. Contours (solid lines) are in hundreds of feet for 200-foot intervals. Isotherms (dashed lines) are for intervals of 5° C.

From the assumptions of the gradient wind equation the divergence of the gradient wind is controlled by three factors: the solenoid term, the latitude effect, and the curvature effect [8]. At low levels, in a homogeneous air mass such as the one discussed here, the solenoid term is negligible. It may be neglected at higher levels when the isotherms are in phase with the contours as in this cold trough at 300 millibars (fig. 14). At low levels ahead of weak troughs the latitude effect predominates over the curvature effect with northward flow producing convergence [4] (see fig. 4). But the latitude effect depends on the first power of the wind speed while the curvative effect depends on the square of the wind speed [7]. Thus, with even slight increase of wind speed with height in the troposphere, the curvature effect becomes relatively more significant with increasing height until eventually it predominates over the latitude effect. The curvature effect in the upper troposphere produces divergence between the trough line and the preceding ridge line [7]. As convergence was found above 500 mb. in this case the latitude effect appears to have been effective in rather a deep layer (see the increased southerly flow south of the rain area in figs. 3, 4 as compared to figs. 1, 2) [10]. Low level convergence and high level divergence produce upward vertical motion as shown by the equation of continuity.

From such reasoning Bjerknes and Holmboe developed a model for the vertical distribution of the divergence ahead of and behind a trough such as the one described in this article (see fig. 1 in reference [11]). Ahead of the trough there is convergence below a level of nondivergence and divergence above. Behind the trough the convergence is above the level of nondivergence and the divergence is below. Empirical studies [7] have found this theoretical distribution of divergence to be a good approximation to the observed.

CONCLUDING REMARKS

Since it is desirable to have an aid in the explanation of weather that occurs without frontal or orographic lift it would seem advisable to indicate such a model as that of Bjerknes and Holmboe on the weather maps. This could be done by transposing the contours of an upper level chart, e. g. the 700-mb. chart, directly upon the surface chart [12]. Currently this is not done for analysis work, however, the 30-hour surface prognostic charts transmitted from the WBAN Analysis Center by facsimile do have the 36-hour 700-mb. prognostics superimposed on them. The spatial model could also be suggested on the surface map by the introduction of a new analysis symbol to represent "trough aloft." The trough aloft superimposed on the surface analysis would bring to mind the possibility of pretrough lifting which, given a convectively unstable air mass, would result in precipitation.

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12. V. J. Oliver and M. B. Oliver, "Meteorological Analysis in the Middle Latitudes", *Compendium of Meteorology*, American Meteorological Society, Boston, 1951. (See p. 721, fig. 2.)

CORRECTION

MONTHLY WEATHER REVIEW, vol. 81, No. 6, June 1953, page 170: Maps in figures 3 and 4 should be interchanged. Map labeled figure 4 is for 1500 GMT, June 8, 1953.

U. S. GOVERNMENT PRINTING OFFICE: 1953